Geometric and Rheological Asperities

in an Exposed Fault Zone

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Abstract

Earthquake dynamics are strongly affected by fault zone structure and fault surface geometry. Here we investigate the interplay of bulk deformation and surface topography using detailed structural analysis of a fault zone near Klamath Falls, Oregon combined with LiDAR measurements of the fault surface. We find that the fault zone has layered damage architecture. Slip primarily occurs inside a 1-20 mm wide band that contains principal slip surfaces with individual widths of ~100 µm. The slip band sits atop a cohesive layer which deforms by granular flow. Several fault strands with total slips of 0.5-150 m also have cohesive layers with thicknesses increasing monotonically with slip. The rate of the thickening decreases with increasing displacement indicating that slip progressively localizes. The main fault is a continuous surface with 10-40 m quasi-elliptical geometrical asperities, i.e., bumps. The bumps reflect variations of the thickness of the granular cohesive layer and can be generated by a boudinage-like instability. As the granular layer is stiffer than its surroundings, the asperities are both geometrical and rheological inhomogenities. Modeling slip along wavy faults shows that slip on a surface with a realistic geometry requires internal yielding of the host rock.
Consistently, our observations suggest that deformation processes in the fault zone include on-going fracturing, slip along secondary faults and particle rotation. Granular flow is an important part of faulting in this locale. Slip surfaces localize on the border of the granular cohesive layer. The on-going slip smoothes the surfaces and thus the structural and geometrical evolution of the granular layer create a preference for the continuation of slip on the same surface. There is a feedback cycle between slip on the surface and the generation of the granular layer that then deforms and controls the locus of future slip.
Introduction

Faults evolve through a complex series of physical and chemical processes acting over seconds to millions of years and on scales of microns to kilometers. Fault zones form with discrete slip surfaces in the fault core (Sibson, 1982; Chester and Logan 1987; Cowan et al., 2003). At the same time, the surrounding rock is damaged and deformed (Shipton and Cowie, 2001; Sibson, 1986). The combination of these processes controls slip (seismic and otherwise) and stress evolution in the fault zone.

One of the major determinants of slip distribution in an earthquake is the presence of asperities (Lay et al., 1982). Although the word “asperity” originally meant a bump on an otherwise smooth surface, seismic studies have inferred asperities from slip distributions. As the seismic data simply define the distribution of slip during the earthquake, the studies raise a fundamental question about whether the origin of the heterogeneities is primarily geometrical or rheological. In this paper we attack the same question but focus on the scale of the slip in a moderate to large earthquake. This is a much finer scale than is accessible through seismic inversions.

Faults have geometric irregularities in the direction of the slip over a large range of scales (Okubo and Aki, 1987; Power et al., 1987; Peacock, 1991). Faults display segmentation, branching, and corrugations which indicate an on-going nucleation and merging of brittle fractures (Stewart and Hancock, 1991; Jackson, 1987; Lee and Bruhn, 1996). Other corrugations such as mullions or boudins occur along large lithological contrasts. For example, John (1987) described low-angle normal faults in Southern California with wavy geometries along hundreds
of meters to kilometers and showed that the amplitude and the wavelength of corrugations vary
with the lithology of the footwall. At the outcrop scale, slip surfaces often have polished
striations generated by abrasion. They can also be affected by adhesive wear, pressure solution,
and growth fibers (Petit, 1987; Means, 1987).

The above picture might suggest that fault geometry is dominated by stochastic processes,
but some measurements indicate that fault roughness evolves with slip. Spectral analysis of
roughness parallel to slip orientation presented by Power and Tullis (1991) can be interpreted as
non-fractal (Ben Zion and Sammis, 2003). Using ground based LiDAR and a laboratory
profilometer, Sagy et al. (2007) demonstrated that slip surfaces of small-slip faults are rougher
than those that slipped larger distances. On-going slip might also generate statistically scale-
dependent roughness. Surfaces of small-slip faults are relatively rough at all measured scales,
whereas those of moderate-slip faults are polished at small scales but contain quasi-elliptical
asperities at scales of a few to several meters (Sagy et al., 2007).

In this paper we address the mechanism that generates this complex surface geometry by
combining detailed structural analysis of the fault structure with measurements of its surface
geometry at a particularly exceptional locale: The Flowers Pit Fault. The fault has a large and
continuous surface exposed over a ~0.5 km hillside. The very fresh surface was unearthed
recently, and there are many exposures of cross sections perpendicular to the fault. Thus the fault
presents a singular opportunity for measuring the relationship between fault surface geometry
and fault zone structure. The fault zone exhibits a consistent damage structure, similar to the
structure of other fault zones in different tectonic regimes (Agosta and Aydin, 2006; Billi et al.,
2003; Chambon et al., 2006). Thus it may provide an example of general faulting processes in the brittle crust.

The geological setting of the Flowers Pit Fault will be described in Section 1. In Section 2 we will discuss the topography of the fault surface with a focus on the 20-40 m scale bumps and the scale-dependence of the accompanying roughness. Section 3 will describe the fault zone architecture starting from the highly localized slip surface and proceeding into the foot wall. This detailed geological work will illustrate that there are distinct layers that exhibit varying degrees and modes of deformation including granular flow and brittle fracture. The thickness of a key layer, termed Layer II, will be tied to the surface topography. Section 4 explores how the fault zone evolves with increasing slip through both direct observations of secondary faults in the Flower Pit complex and two separate modeling exercises. The combined results of Section 4 show that the internal deformation processes are involved in the generation, evolution, and destruction of the asperities. In this fault zone, we will find that asperities are simultaneously geometric and rheological features. They are manifestations of damage evolution in the fault zone. There are feedback processes that generate internal damage due to slip on the fault and then control the distribution of future slip through heterogeneities in the damage structure.

1. Geological Setting

The Flowers Pit Fault is a young normal fault located southwest of Klamath Falls, Oregon (Fig. 1a-b). The fault belongs to the Klamath Graben Fault system in the northwest province of the Basin and Range (Personius et al., 2003). Quaternary and Holocene activity was found on several nearby faults from the same system (Bacon et al., 1999), and the area is seismically active, including a sequence of magnitude 6 earthquakes in 1993 (Braunmiller et al., 1995). The
fault itself is exposed by recent quarrying, and thus the surface is fresh (Fig. 1c). The exposed footwall contains mostly andesite sequences. Fine-grained sandstone and arkose layers with thicknesses of a few meters appear above the wide, faulted andesite layers. These lake deposits are horizontally layered and covered by newer andesite and basaltic flows. The uppermost exposed sediments are gravels that are unconformably bedded on the andesite and on the sandstone.

The fault strikes to the WNW and creates a zigzag pattern (Fig. 1b). The displacement along the fault is localized along a relatively continuous fault surface. The fault is exposed along 550 m and includes three large surface exposures with a composite area of ~6000 m$^2$. We assume that the total displacement is 100-200 m based on the appearance of patches of sediments and coarse gravels on the hanging wall vertically diverted to about 100 m below their original locations on top of the hill.

The current quarrying exposed three separated sections. The northwestern one dips 55°/242° (dip/dip direction) on its northwestern edge and curves to 57°/234° on the southern end. This segment is exposed continuously along 150 m without any major splitting or branching. The longest exposed sections in the direction of the slip orientation are 30 m long. Branches and secondary fractures are rarely observed. The middle segment dips 66°/204°. It is exposed along a relatively small zone, but includes areas of up to 37 m of continual exposure parallel to the slip direction (Figs. 2a-b). Ongoing quarrying in this section also provides very good exposure of the fault zone. The southeastern segment dips 64°/214°. This section is more eroded, and therefore fewer chunks of fresh fault surface are exposed. However, because of the erosion, this section
provides many opportunities to observe the structure of the fault zone under the surface. The southeastern segment is bifurcated on its northeastern side into two separate segments (Fig. 1b-c). One segment has a total displacement <50 m and strikes NNW (Fig. 1b), while the other is covered by gravels, but probably merges with the middle section.

The slip on all the segments is oblique and the orientation of the last slip is measured using tool marks and striations. We also use our dense LiDAR measurements to follow the orientation of larger geometrical irregularities such as bumps and depressions (Fig. 2a and Appendix A). We found that the orientation of small scale marks and larger undulations are similar. The slip trend on the segments is 246°-271°. The slip sense and dip direction can differ by as much as 38°, as occurs on the middle segment.

2. Fault surface geometry and roughness

The most striking geometrical features on the surface of The Flowers Pit fault are quasi-elliptical bumps (Figs. 2). The typical bump shape is elongated with the major axis of the bump parallel to the slip orientation as recorded by small scale striations and abrasional tool marks.

The exposed bumps are relatively long, and most are only partly preserved. Several examples on the Flowers Pit Fault demonstrate that the bumps are exposed as both protrusions and depressions (Fig. 2). Such quasi-elliptical structures are observed on other faults in the western U.S. (Sagy et al., 2007) and in Italy (Fig. 2d) and may represent a common structure.

Fig. 3a presents two examples of bumps from the northern and the middle segments of Flowers Pit Fault. Next to each image are profiles extracted from the LiDAR data parallel to the slip taken nearly along the long axis of the bump. The major and minor axes of the bump in the
northern part of the fault are 20 and 5 m long and the maximum height relative to the margins is 30 cm. This bump is the best-preserved of at least 5 large-scale bumps that have been identified on the northern part of the fault (See Sagy et al., 2007 Fig. 4b). A less symmetric bump is presented on the right side of Fig. 3a with major axis length of 15 m and maximum amplitude of 55 cm. In general, bumpy surfaces appear along the entire surface in Flowers Pit Fault. However in the northwestern segment their dimensions are more constant and very similar to the one that is presented on the left side of Fig. 3a. The other two segments contain zones with smaller and larger bumps.

The bumps result in wavy profiles in the slip-parallel direction. Along the long axis of the bumps, the waviness has an amplitude of dozens of centimeters at the scale of 10+ meters that defines the large bumps. Below this scale, smaller wavelength bumps are generally a few millimeters high or less (Fig. 3a).

This observation of distinct behavior at different scales is generalized in Fig. 4b. The power spectral density is calculated for different sections of the fault as a function of the wavelength. Power spectral density is closely related to the RMS roughness (Brown and Scholz., 1985) and is calculated by applying a multitaper Fourier transform to individual profiles (Appendix A). Here we present the spectra based on the data collected by both the field scanner (top profiles) and a laboratory profilometer (bottom profiles). Every curve is the average of the spectra from 200-600 profiles. For example, the blue and the red curves in Fig. 3b are calculated from the section observed in Fig. 3a (the bumps). In general, the slopes of the power spectral densities of the field scanner data are steep above wavelengths of ~1 m, but become moderate at shorter wavelengths. These variations in the curves’ slope indicate that below the wavelength of the
large bumps, the surfaces are so smooth that most of the topography variations are below the resolution level of the field scanner (<3 mm). Moreover, examination of sections with large bumps (Fig. 2a-b) suggests that because the surface is so smooth, part of the measured roughness in the LiDAR scans is contributed by fractures and by small-scale erosional features which are not necessarily related to the faulting.

To measure the surface roughness at scales that are smaller than can be detected by the field scanner we used a laboratory profilometer. The roughness of two hand samples with relatively fresh surfaces was measured along lengths of 0.1 to 70 mm, and their power spectral density is shown by the lower curves in Fig. 3b. The moderate slope of the power spectral density curves corresponds to relatively weak variations of roughness height as a function of their wavelength (close to logarithmic decay of the RMS roughness). This moderate change of roughness is mainly an expression of abrasional processes that are involved in the generation of the uppermost part of the surface (Layer I, Figs. 2a). We also note that preferred erosion of Layer I, which is less cohesive than the layer below (Layer II), also contributes to the small scale roughness even in relatively fresh samples.

To summarize, elongated elliptical bumps with lengths of dozens of meters are present on the surfaces. At wavelengths shorter than the bumps’ scale, the surface is smooth as with a logarithmic decay of the RMS height with wavelength. Abrasive wear might be the main mechanism controlling the small scale roughness as discussed by Sagy et al. (2007). These observations differ from roughness measurements of many natural joints and faults that suggest self-affine roughness (Brown and Scholz., 1985; Renard et al., 2006). In the following sections
we present a detailed structural analysis of the Flowers Pit fault zone and discuss the inter-
relationship between the surface evolution and the fault zone structure.

3. The structure of the fault zone

To investigate the interaction between the fault geometry and the evolution of the internal
damage we present a detailed analysis of the fault zone structure. We show that the fault zone
displays distinctive damage characteristics that are relatively continuous parallel to the fault
surface, but abruptly change as a function of the distance from the surface. Thus, we use the term
“layer” to describe the damage architecture (Fig. 4). The hanging wall is exposed only in small
patches and therefore it will not be discussed in detail.

3.1 The band of principal slip surfaces (Layer I)

The uppermost damage zone (adjacent to the surface) includes evidence for extreme
localization on multiple slip surfaces accompanied by abrasional, fluid injection and granular
flow structures.

A thin reddish layer is exposed at the fault surface (Fig. 5a). This zone typically has a width
of 0.1-5 mm and occasionally reaches a width of 2 cm. The layer is exposed in very fresh zones
on the fault. In more eroded areas, the layer fills elongated striations and indicates abrasive wear
(Petit, 1987). Examination of samples from this zone under a microscope reveals that the region
typically contains one to three slip surfaces which are expressed by linear fine-grained bands
inside altered granular material (Figs. 5b-d). The mineralogy of this layer contains 90%
plagioclase. However, the reddish color in this layer (Fig. 5a) is contributed by hematite as
determined by X-ray diffraction (XRD).
The slip surfaces are parallel to the primary surface and are distinguished from the surroundings by particle size. Typical slip surface thickness is 100-500 µm (Fig. 5c). Using SEM analysis, we found that the sizes of the fragments inside a single slip band range between 10 µm-0.1 µm (Fig. 5b). The small particles are mostly shattered feldspar crystals and are much less rounded than the larger particles between the bands. As the size, shape and color of grains varies between the slip surfaces, we infer that several slip episodes are recorded.

In many cases, we observed secondary fractures that branch from the slip surface (Fig. 5c). Interestingly, we found that the granular injections penetrated only to one side of the fault (toward the footwall). The fractures tend to fork at high angle to the surface and then rotate to 50°-70° degrees from the main slip surface. They are filled by the same fine-grained material as the main band. The observed branches extend no more than a few millimeters. The branches inject the fine-grained material into the layer below (Layer II), thus sometimes creating islands of fine-grained material (Fig. 5c) in a sea of coarser-grained material. Similar secondary cracks are also observed at a larger scale (Fig. 5a) where the fine-grained material of the principal slip zone is injected a few centimeters into the layer below.

These branches are evidently opening-mode fractures because they do not shear the slip surfaces; they inject fine-grained material and contain secondary veins (Fig. 5d). The veins are evidence of the existence of fluids during slip. The overall picture strongly suggests that the branches are hydrofractures which indicate internal pressurization of the slip surface (Byerlee, 1990; Rice, 1992; Brodsky and Kanamori, 2001). The association with injected material suggests increased internal pressure. When the internal pressure exceeds the confining pressure by the
tensile strength of the rock, a hydrofracture might be generated perpendicular to the main compressive stresses. In this case, the granular material has no tensile strength and the fault is near-surface, so only moderate pressures may be needed. Hydrodynamic lubrication with slurry of viscosity of 10 Pa-s in the observed 100 µm wide bands with thickness variations of 0.1% will result in a pressure increase of 50 MPa due to the shearing of the fluid over 1 m of slip at a typical earthquake slip velocity of 1 m/s (Brodsky and Kanamori, 2001). This overpressure is sufficient to drive the injections at depths up to nearly 2 km.

Fluidization and ductile deformation are also documented in this band. A well-preserved section of Layer I was found above depressions or in relatively planar zones of the slip surface. In a few locations, the layer reached a thickness of ~2 cm. Fig. 5e-f shows a thin-section from this thick region. The upper part of the section (Fig. 5e) shows planar sub-parallel slip bands that shear granular medium from Layer II. On the lower part (Fig. 5f) however, wavy layering and boudinage structures appear in some of the layers, thus indicating ductile deformation of the layered medium and fluidization of the granular material. Similar structures are observed during experiments of shear of rocks under dynamic friction conditions (Boutareaud et al., 2008).

3.2 The cohesive granular layer (II)

The layer below the uppermost slip band is the most distinctive layer in the fault zone. It contains cohesive rock that is persistent over the entire exposure (Figs. 1c & 3a). The layer is exposed under Layer I or in places where Layer I is eroded. In the middle and the southern segments the layer is 5-100 cm thick, however, in most of the exposure it is about 10-20 cm. As will be shown below, the layer provides evidence for continuous internal deformation by
granular flow, fracturing and slip. “Granular flow” is a rheological term for relative rotation of solid particles during bulk deformation (Campbell, 1993). This internal rotation of particles distinguishes granular flows from more commonly studied rheologies like elasticity or Newtonian viscosity.

Thin-sections show that the layer contains aggregates of crystals comprised mainly of plagioclase rotated inside a matrix of finer particles that includes single crystals or fragments of plagioclase (Fig. 5c). The grains generally have dimensions of a few millimeters to a few centimeters. Extremely small-scale grains (diameters <1 µm) appear between the larger grains (Fig. 5c) as observed on other faults (Engelder, 1974). The roundness of the grains in this layer (Fig. 5 c, e.) suggests that significant particle rotation occurred during faulting.

One of the mechanisms to bring small grains into the layer is the injection of fine grained material from the principal slip surfaces (Fig. 5a,c-d). Such injections of fluids mixed with ultrafine grains contribute to cementation as evidenced by the vein in Fig. 5d. Interestingly, we found no clear correlation between particle size and distance from the surface inside the layer. In most cases, a few millimeters from the surface, grains’ radii can achieve 1-3 cm, which is the same order of magnitude as grains that are a few centimeters from the surface. Consequently, we infer that the cohesiveness of the layer is a result of the on-going granular deformation, which increases the adhesive forces between comminuted grains as the surface area increases (Gilbert et al., 1991), repacks the rock volume to a denser configuration (Aydin, 1978), and promotes fluid transport into the layer (Fig 5d). The combination of effects lithified the granular material.

X-ray diffraction analysis of this region suggests no evidence of cementation beyond that represented by the injection features of Fig. 5d.
Brittle deformation at a larger scale is also observed in the layer. Joints that cross the entire layer are likely connected to the unloading of this zone during the on-going faulting. Small normal faults with displacements of a few millimeters appear mostly at acute angles to the main surface. In most cases, these faults have a sense that is sub-parallel to the slip orientation recorded by the striations and they typically cross the entire layer creating an S-shape in cross section. Continuous volumetric deformation of the layer is evidenced by small faults with spacing of a few centimeters that fragment the layer. Such faults indicate re-granulation of the already cohesive layer.

The surface undulations discussed in the fault geometry and roughness section are related to abrupt variations in the Layer II thickness. The layer widens under fault surface protrusions. For example, the maximum height of the bump in the right side of Fig. 6a is 45 cm relative to the nearest depression on its right side. The maximum width of Layer II there is 110 cm. Under the depression on the left side, the minimum width of the layer is 5 cm. Figure 6b presents two more examples of the thickening of Layer II under the bumps and thinning of the layer under depressions. The maximum exposed thickness of Layer II in Fig. 6b (top, left) is 0.6 m, while the maximum amplitude compared to the nearest depression is 0.35 m. Under the nearest depression (Fig. 6b top, right), the width of the cohesive layer is again 5-10 cm. In all examples exposed, we found that the cohesive layers thicken under the bumps (Fig. 6c). Following this observation, we infer that the wavy appearance of the surface is due to variations of thickness of Layer II.

Fig. 7 provides evidences that suggests that bumps are brittly deformed and destroyed during slip as the geometry of the slip surface evolves. In contrast to the general correlation between bump appearance and the thickening of Layer II (Fig. 6c), Fig. 7a has a relatively thick
Layer II with no clear bumpy appearance. Secondary normal faults that cross the layer at a sharp angle to the main surface appear in the photo. Fig. 7b shows similar normal faults that are associated with extension of the layer parallel to the slip. The displacement on these faults separates the entire bumpy structure into rotated blocks crossed by the present main surface.

The cartoons in Fig. 7 present our interpretation of the observation. Following our previous observations in Fig. 6, the relatively thick Layer II in figure 7a suggests that once it was a bump, but with continuation of slip the blocks heads was worn and truncated to generate the smooth current main slip surface. Such a domino-like collapse is typical for the compression of a competent layer inside a less competent medium when the competent layer becomes extremely stiff or the amplitude of the bumps becomes extremely large (Johnson and Fletcher, 1994; Goscombe et al., 2004). Thus, although Layer II is more cohesive than the surrounding damage zone (the hanging wall and Layer III), it still deforms with increasing slip and the current bumps shown in Figs. 2 and 6 are probably temporary.

### 3.3 The non-cohesive fragmented layer (III)

The damage zone under Layer II has two main characteristics. First, it is much less cohesive and fragments easily. Second, it has deformation that destroys the original mesostructures of the andesite (beds, veins and fractures). However, in contrast to the observations in Layer II, the microstructure is generally well-preserved (Fig. 8).

The layer is 0-2 m wide. The width variability is probably due to internal shear flow and stretching. We found no correlation between the Layer III width and the change in thickness of Layer II or the bumpiness of the fault surface. Particle diameters are between a few millimeters
and 30 cm. In some cases, several secondary faults from Layer II penetrate dozens of centimeters into this layer and then fade away (Fig. 7). Under a bump, Layer III is always separated from Layer II by a fault (Fig 7). Elsewhere, when such a fault is absent, the transition from cohesive to non-cohesive material occurs over a few centimeters.

The lithology in Layer III contains 90% plagioclase feldspars like the host rock and the cohesive zone. Structurally, however, the fragments differ from those in Layer II. Thin sections show that the particles contain opening fractures and shear bands at spacing of a few millimeters to a few centimeters (Fig. 8). The shear bands contain high concentrations of small faults. Many of them appear in conjugate geometry. The particles inside the bands include sheared plagioclase minerals and their grain sizes are as small as <1 µm. There is no preferred orientation for the small scale shear bands or for the tensile fractures (Fig. 8). Outside of the shear bands, unfractured plagioclase crystals typically have 50-200 µm long axes and the original magmatic structures are usually preserved (Fig. 8). The crystals are chemically altered along fractures and sometimes inside the fragments.

Interestingly, the hanging wall remnants show very similar structure and bulk deformation as Layer III. If such non-cohesive material also borders the main slip surface (Layer I) from the hanging wall side, it can explain the preservation of the bumpy topography of the cohesive zone as it can deform easily to accommodate the changes in geometry. In many places in the footwall, Layer III is separated from Layer II by a fault which might have been the main fault in the past (Fig. 7). The observations suggest that a symmetric sandwich could be developed with Layer II in the middle bordered by slip bands (Layer I) and Layer III on each side. Yet, in most of the
observations in the Flowers Pit Fault only one slip surface is dominant while the other has been abandoned.

The above observations of fracture and shear suggest that the non-cohesive Layer III is a product of faulting. In addition, less competent flows in the andesite sequence might also contribute to the development of the non-cohesive zone. Such layers tend to stretch and flow along the fault. However, we found that far from the fault all of the exposed andesite flows have large fragments and are cohesive. Moreover, the non-cohesive damage zone is not unique to Flowers Pit or to the specific lithology. Similar damage zones were observed in other normal faults that occur in different tectonic and lithologic environments (Chambon, et al. 2006).

In summary, the observed microscopic and macroscopic deformation in Layer III includes fracturing and shear in its entire volume, however the amount of rotation and comminution are small compared to Layer II. One way to explain these observations is by breakages due to mismatch of the fault surfaces during sliding (Sibson, 1986), combined with smearing of non-competent andesite flows parallel to the slip surface (Aydin and Eyal, 2002).

3.4 More Distant Damage

Further from the fault surface than Layer III, we observe the original macroscopic features of the host rock such as flows and layers, although the rocks are still highly fragmented. The fracture spacing and character gradually changes. Layer IV is a zone in which small faults and joints appear with more coherent structure at the mesoscale. Intensity and fragmentation of fractures is more influenced by the width and the rheology of the host andesite layers here. Layer V is defined as the zone which is dominated by joint sets which appears perpendicular to the
andesite layers, although fragmented zones still exist. Exposure limitations prevent measurement further than ~50 m from the fault.

Figure 9 presents the spacing of fractures in the footwall as a function of the distance from the fault. All distances were projected to be normal to the exposed fault surface. The data include measurements from Layer III-V. Most of the data was collected by mapping fractures with a gridded frame (Sagy et al., 2003). We also used photographs to map the fracture density. In Layer III, fracture spacing was measured by photographs and by thin sections like the one in Fig. 7.

The most striking observation in Fig. 9 is the localization of damage near the fault and the non-linear decay of the fracture density. The measurements in the first meter are sampling fractures density in Layer III and its local environment. The cumulative offset that repeatedly deforms this layer results in a very dense fracture spacing. Changes in the fracture density further from the slip surface are relatively small. Beyond Layer III, the fracture density falls off as approximately the square root of distance. The slope of the fracture density curve suggests that up to at least ~ 50 m from the fault surface fracture density is still influenced by the fault. We did not manage to measure the background fracture density directly.

About 10-30 m and further from the fault (layer V), most of the observed fractures are opening-mode joints. Some of them are sub-parallel to the main fault segments. They cross layers, are very open when exposed and can reach dozens of meters in height. They are probably related to the extension of the footwall near the fault (Sagy et al., 2003). However, the spacing of these joints is much smaller than the average fracture density which further reinforces the conclusion that the fault is still affecting the damage at the most distant measurements in Fig. 9.
4. Structural evolution of the fault damage and fault surface

4.1 The development of the cohesive layer with slip

The cohesive layer (Layer II) adjacent to the slip surface is not unique to the main fault. Similar layers also appear in branches of the main fault and in secondary faults in the fault zone. We showed that one of the principal characteristics of this layer is that grains are shaped by rotation and comminution (Fig. 5). In this respect, the layer differs from the damage zone further from the fault surface (Layers III-V) which is mostly fractured and fragmented. It is also different from the slip band zone (Layer I). Experiments suggest that such layering architecture is a typical outcome of deformation in granular material (Mueth et al., 2000) and strongly dependent on the roughness (Chambon et al., 2006). Following this observation we deduce that Layer II is closely related to wear material that is generated in shear experiments (Kato and Adachi, 2000).

Fig. 10 demonstrates the relationship between fault displacement and thickness of the cohesive layer (Layer II) in six individual faults in the Flowers Pit fault zone. The measured faults are the main fault, the fault on the other side of the horst (Fig. 1) and four smaller faults. We measured the displacement of these faults using the layer offsets. Thickness of the cohesive layer was measured in 5-40 places along each fault, depending on the exposure. As the data was collected only in the Flowers Pit Fault zone, it represents similar rocks that deformed under similar physical and chemical conditions. We found that the cohesive layer is generated in an early stage of the fault and continues to evolve (Fig. 10). However, on very small faults (displacement <0.5 m) the layer could not be identified.
Our observations suggest a monotonic, non-linear positive correlation between the displacement of the fault and the thickness of the cohesive granular layer. Fig. 10 also suggests that the spatial gradient of the thickening (dT/dx) decreases as a function of the displacement (x), i.e., d²T/dx²<0. The trend is consistent with the decrease of asperity amplitude during fault evolution that we measured in our previous study that compared the roughness of faults with differing slips (Sagy et al., 2007). The decrease of the thickening with slip is comparable to the exponential decay of the wear production with increasing slip during the asperity-breaking stage (or running-in wear stage) of laboratory experiments (Quineer, 1965; Wang and Scholz, 1994; Wong et al., 1998). However, in these experiments the length of the slip surface is constant, but the dimension of a natural fault is a function of the total offset on the fault (Scholz, 2002).

The decrease of the layer production with slip is evidence of the localization and maturation processes during fault slip. Although variations in the layer thickness still exist as evidenced by the bumps (Figs. 2, 6), the slip on average is progressively localized as a function of the displacement.

### 4.2 Internal failure during slip on wavy faults with bumps

The instantaneous fault shape imposes important boundary conditions for the next potential event regardless of the processes that generated the geometry. The Flowers Pit Fault zone suggests a view of fault zones as dominated by a wavy localized main fault surface bordered by deformed and damaged rock. Let us now investigate the constraints on an earthquake that could occur on an active fault like the Flower Pit exposure. If the fault is loaded during the inter-seismic period and responds by slip along a pre-existing surface during an earthquake, we can
calculate the stresses around such a fault before the host rock yields. We will use this calculation
to test how much a wavy fault can slip before breaking the host rock and predict where this
yielding occurs.

Although the 3D topography of the fault is characterized by elliptical geometry, parallel to
slip the fault is wavy. Simplification of the geometry in Flowers Pit may allow us to answer the
above questions by calculating the stresses near a 2D wavy fault (Saucier et al., 1992; Johnson
and Fletcher, 1994; Chester and Chester, 2000). In this analytical approach the stresses are
calculated by adding the contribution of small-scale sinusoidal geometrical perturbations to a
planar fault.

In the present analysis, we use the model of Chester and Chester (2000) to calculate the
stresses around a frictional wavy fault surface hosted by an elastic medium (Appendix B). The
calculation is performed on a 60° dipping normal fault at depth of 1 km. We used our
topography measurements (Fig. 3) to define a typical aspect ratio of asperities: $H/L = 5 \times 10^{-3}$
where $H$ is the asperity height and $L$ is the length (Fig. 11a). The host rock yields when the
stress in the rock exceeds the local Coulomb failure threshold (Eq. B4).

Fig. 11a presents the values of the main stress component $\sigma_1$, near the fault. The stresses
calculated here are for zero displacement under stress relation $\sigma_H/\sigma_V=0.5$. This relation is
realistic for normal fault regimes (Reches et al., 1992). However, because we assume that
stresses in this location are not necessarily in a critical condition, the imposed stresses are below
the condition of Coulomb failure of the host rock. The figure shows that the geometry generates
a stress perturbation and rotation near the fault. As is expected, the value of the maximum
principal stress ($\sigma_1$) increases with respect to the vertical stress in the front of the bump and decreases in the lee of the bump. Also, the influence of the geometrical undulations on stress perturbation is limited. At a distance of more than 10% of the wavelength, the undulations are negligible (Fig. 11a). The host rock is predicted to yield in the center of the bump with a displacement of only 0.2% of the wavelength of the fault surface roughness.

The initial yielding of the host rock as a function of the slip distance, tectonic stress field and roughness is calculated using this model (Fig. 11b). Smooth surfaces are predicted to absorb larger amounts of slip before internal yielding. This model also predicts that slip along a fault surface with realistic geometry requires internal yielding of the host rock. For example, slip of about 1-10 m along surface without damaging the host rock requires a non-realistic amplitude-wavelength ratio of 1:10^6. This prediction suggests that earthquakes involve deformation of the host rock. This deformation can be an important source for energy dissipation and a contributor to earthquake arrest.

4.3 The Formation of the Bumps

We observe that the fault surface bumps are manifestations of lenses of cohesive granular material (Fig. 6). In places where Layer II is thick, the slip surface heaves upwards and vice-versa. Since the dominant characteristic of Layer II is the significant internal deformation, it is natural to look for a flow explanation for the thick lenses.

Brittle and ductile shear in layered rocks with significant rheological differences between them are typically involved with boudinage and mullions (Smith, 1977; Goscombe et al., 2004). Fig. 5f presents a ductile micro-boudinage in Layer I, while Fig. 7 demonstrates that boudinaging
might be an important mechanism in the brittle deformation of Layer II. Here we pursue a third type of boudinages in the thickened granular flow. Boudinage structures form in a competent layer under compression when it is embedded in a softer material (Johnson and Fletcher, 1994; Smith, 1977; Twiss and Moores, 1992). There is also a closely related instability between layers of contrasting competence under simple shear rather than pure shear that has been previously unexplored in the fault zone literature. A straightforward extension of previous boudinage instability work shows that for positive shear stress, thickness instabilities form in the more competent layer (Smith, 1977) (Appendix C). Thus, cohesive Layer II is likely to distend due to the internal flow.

The growth rate of the thickness instabilities is strongly wavelength-dependent. For any given set of parameters, there is a wavelength that has the maximum growth rate. This wavelength then dominates the flow and its preserved structure. Therefore, to explore the possibility of flow instability, we measure the typical wavelengths of the bumps and compare them to the computed instability wavelengths for a linearized shear flow model with typical parameters (Appendix C). The rheology of the granular flow at the appropriate conditions of fault motion is extremely uncertain (Brodsky and Kanamori, 2001; Lu et al., 2007). We therefore employ a non-Newtonian flow model to explore a range of possible behaviors by employing the general non-Newtonian constitutive law of $\sigma^n \propto \dot{\varepsilon}$. If $n=1$, the flow is Newtonian, while a large value of $n$, such as 10, implies nearly purely plastic behavior. To model the strength contrast observed in Fig. 4, the cohesive granular layer (Layer II) is modeled with a higher effective viscosity than the surroundings.
The wavelength of the bumps in the slip direction is \( \sim 10\text{-}40 \text{ m} \) and the thickness of the layer is typically 0.1-0.2 m, so the ratio is \( \sim 50\text{-}400 \). Fig. 11 shows a successful match between the model for a range of effective viscosity ratios \((m)\) of \( 10^{-4}\text{-}10^{-6} \) and a variety of values of the stress exponent \( n \). This range is generous. It appears relatively easy to produce the observed bumps through a flow instability in the granulated layer.

The preferred wavelength of the instability explains the distinct difference in spectrum between the sub-bump scale wavelengths and the asperity scale in Fig. 6. The bumps are distinct features and shorter wavelengths are smoothed out by the bulk flow of grains combined with abrasion.

As slip progresses, the granulated layer thickens and the wavelength of the bumps increases. As shown by Fig. 10, the rate of thickening with increasing slip decreases, so we predict that the wavelengths will lengthen with slip at a decreasing rate as well. This prediction provides a useful test of the hypothesis of a flow origin for bump.

Clearly much more work is needed to fully validate this model for bump formation. As indicated above, if the bumps are generated by shear zone instability, the thickness of the granular layer should co-vary closely with the wavelength of the bumps. Bumps on fault zones of various slip should be measured and their wavelengths compared to the local thickness of the granulated layer. The fault zone layer deformation should also be systematically measured in the laboratory and the appropriate rheologies used in an improved model. Such work is beyond the scope of this primarily observational paper. For now, we confine ourselves to presenting a plausible model for the observations and suggesting the above tests for future work.
4.4 Localization of slip

One of the most robust results of the field study is that the principal slip surfaces are on the edges of the cohesive granular zone. Although there is some distributed deformation throughout the fault zone as illustrated by the granular flow textures (Fig. 5c) and the secondary faults in Layer II (Fig. 7), the principal slip surface is clearly distinguished (See sect. 3.1). For instance, only on this surface do we see clear abrasional striations, the reddish hematite coating and the micron-scale fragmented texture of Fig. 5b.

There are at least two mechanisms leading to the slip concentration at the granular flow boundary. First of all, an interface between dissimilar materials commonly results in localization due to the difference in compliances. An imposed stress field results in different strains in the two media and if the difference is large enough offset results (Kellie and Tyson, 1965).

Secondly, granular flows have particularly pronounced boundary localization. The phenomenon motivates the usual practice of introducing sand paper into experimental configurations in order to generate internal deformations (Campbell, 1993). Lu et al. (2007) observed in shear experiments that sand flows tended to localize slip into a region that extended ~2 grain diameters from the wall of the shear cell. Campbell (1993) explains the localization as a result of the inability of a relatively smooth wall to transmit angular momentum. Since the curvature of the wall is much less than that of individual grains, movement of the wall does not start grains spinning relative to each other. The lack of relative rotation results in the grains being locked together and simply sliding past the wall. In other words, near the wall, the frictional sliding threshold for individual grains is reached more easily than the moment threshold for rotation.
Far from the wall, the interlocking grains effectively transfer angular momentum and relative rotation occurs in preference to sliding. The implication for natural faults is that the on-going smoothing of the main surface and the generation of the stiff cohesive layer act together to continue localization and slip. As the granular layer develops from wear particles, the slip remains at the boundary. Progressively, the boundary itself becomes smoother, abrasion is reduced and the rate of the granular layer growth decreases.

4.5 The cycle of internal deformation and slip surface generation

It has been previously recognized that fault topography can affect internal deformation off-fault (Kim et al., 2004). We have shown in this study that internal deformation also affects fault topography to form a feedback cycle. Slip on the fault surface produces wear particles that form cohesive granulated layer (Layer II) forms adjacent to the slip surface. Continued slip injects new particles into the granular layer. The granular flow deforms and has regions of variable thickness. These variations of thickness are accommodated by internal deformation in the non-cohesive Layer III. The ponding of the grains may be governed by flow instabilities, akin to boudinages. The slip surface localizes on the edge of the cohesive granular layer because of the contrasting rheologies and the granular flow behavior. Since the primary slip occurs at the boundary, the thickness variations of the granular layer generate fault topography. The fault topography then controls future slip events and hence future internal deformation. Individual fault bumps do not persist over the long-term history of the fault, as illustrated by the preserved remnants of truncated bumps and the calculations of the internal failure
conditions in Fig. 10. Instead, bumps are dissected as new fault surfaces form. It is likely that mature, thickened Layer II’s eventually become too stiff to deform internally leading to their brittle truncation. New bumps are also continually being generated by the internal deformation of the granular layer.

This feedback cycle of deformation and localization is profoundly different from the traditional view of fault zones as simple frictional contacts (Brace and Byerlee, 1966; Scholz, 2002). It suggests that a complete model of either energy dissipation or stress accumulation must account for both the internal and surface deformation. Many of these processes have been included as pieces of the energy budget and fault dynamics, but their intimate relationships have not been recognized. The grinding and granular flow feeds the granular flow that in turn controls slip localization.

5. Conclusions and Implications

We showed here that slip surface geometry is strongly connected to the fault zone architecture. The Flowers Pit Fault has a continuous surface with elongated quasi-elliptical asperities (Fig. 2). The geometry of the fault surface is correlated with variations in thickness of the granular cohesive layer (Layer II). A protrusion of the surface indicates a thickening of the layer below it (Fig. 6). As the granular layer is stiffer than its surroundings, the asperities simultaneously serve as both geometrical and rheological inhomogenities. Asperity generation is driven by the variation of rigidity during the evolving Layer II and Layer III. Abrasive wear on the surface (Layer I) and bulk granular flow in Layer II are acting together to generate a surface which is smooth at small scales but bumpy at scales greater than a few meters.
The modeling of wavy faults demonstrates that the evolution of the fault in general and the cohesive layer in particular, involves deformation far from the slip surface. The modeling is supported by the field observations (Sect.3) that indicate three main deformation mechanisms in the fault zone: rotation of grains, fracturing, and slip. Thus, granular flow with considerable rotation of particles is a major mechanism for creating the observed structure of the fault zone (Figs. 4, 7). It is important to note that the host rock itself is not granular in origin and therefore granulation by fracturing and comminution occur during the faulting.

The slip along small faults and the rotation of grains can be significant contributors to the dissipation of energy during slip events, in addition to heat generation by slip along the main surface and surface energy utilized in fracturing (Wilson et al., 2005). However, our observations (Fig. 10) also suggest that the rate of wear production decreases as a function of slip. These observations together with the roughness measurements (Fig. 3) indicate that less energy will dissipate in the fault zone as the fault mature. Thus, for a given amount of stress, slip along a mature fault is likely to be larger than slip on a relatively less mature fault.

Even though deformation occurs in the bulk, the major slip surface of the fault tends to localize in 100 micron bands embedded in a 1-20 mm wide zone on one or both sides of the cohesive layer. The main fault is confined on the borders of the layer although both sides of the layer (the hanging wall and layer III) are weaker (See also Chambon et al., 2006). Such localization of deformation is well documented in complex materials as the two dissimilar materials have different compliance and thus different strains as a result of the imposed stresses (Kellie and Tyson, 1965). The implication of this observation for natural faults is that the on-
going smoothing of the main surfaces and the generation of the stiff cohesive layer act together
to continue localization and slip.

Therefore we suggest that fault structural and geometrical both evolve with increasing slip.

A cycle of damage and internal deformation ties the fault surface and layered architecture
together. Together, the geometry and structure conspire to localize the fault. Together, they result
in the continuation of slip within an extremely narrow zone.

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Appendix A

For measuring the geometry and the roughness of the fault surface over scales of 3 mm – 500 m, we use the Leica HDS3000 ground based LiDAR (Light Detection and Ranging) tool. The scanner can aim its laser beam in a wide range and produce high-density measurements. The LiDAR has several advantages for our research compared to any other method. The accuracy of the measurements enables a reliable quantification of the data. The scanner also enables measuring of structures that are technically hard to approach, such as the middle segment of the Flowers Pit Fault (Fig. 2a). Data that are collected from different locations are integrated to create a 3D picture of large areas. The main advantage of the instrument for roughness measurements is, however, the number of the points sampled. A typical point cloud from a single scan can includes millions of points which can be interpolated to generate a topographic map or thousands of profiles (Fig. 1, 2). Thus, the technique allows a statistical approach when calculating roughness. The Flowers Pit Fault was scanned during two separate expeditions. The first focused on scanning discrete parts of the fault surface from different orientation and at different resolutions, from 3 mm to 2 cm, while the second was dedicated to continuous measurement of the fault surface at a resolution of 1 cm (Fig. 1d). High resolution measurements with point spacing of 3 mm were scanned from 20-40 m. All scanning expeditions included measuring a reference board which included small cubes with known heights of 3 mm, 6 mm,
and 11 mm. We use this reference to compare between the separate measurements and to define
the accuracy of the measurement after the analysis. Measurements were then combined to create
a complete picture of the fault geometry.

The power spectral density is a convenient measure of roughness as a function of
wavelength as it is directly related to the RMS roughness (Brown and Scholz., 1985). We
calculate power spectral densities using a multitaper Fourier Transform method on each profile
individually and averaging 200-600 spectra per curve.

Appendix B

For calculating stresses along a wavy fault we follow Eq. 7a-c of Chester and Chester
(2000). In their model, Chester and Chester performed a linear perturbation analysis of the
stresses along a frictional discontinuity in an elastic homogeneous medium. Stresses are
computed by adding the stresses that are contributed from small sinusoidal perturbations of a
frictional surface to those which are generated by the planar surface. The governing equations
for a fault surface perturbed around the z=0 plane are

\[
\sigma_{xx} = \bar{\sigma}_{xx} + Ae^{-iz} \left( \frac{UEI}{4(1-\nu^2)} (-1 + lz) \cos(lx) + \left[ \bar{\sigma}_{zz} \left( 1 - k + 2\mu^2 \right) - \frac{\mu U E I}{4(1-\nu^2)} \right] (lz) \sin(lx) \right)
\]

(B1)

\[
\sigma_{zz} = \bar{\sigma}_{zz} + Ae^{-iz} \left( \frac{UEI}{4(1-\nu^2)} (-1 - lz) \cos(lx) + \left[ \bar{\sigma}_{zz} \left( 1 - k + 2\mu^2 \right) - \frac{\mu U E I}{4(1-\nu^2)} \right] (lz) \sin(lx) \right)
\]  

(B2)
\[ \sigma_{xz} = \bar{\sigma}_{xz} + A e^{-lz} \left( \frac{UEI}{4(1-v^2)} (-lz) \sin(lx) + \frac{\mu U EI}{4(1-v^2)} (1-k^2 + 2 \mu^2) \right) \]

(B3)

where \( \bar{\sigma}_{ij} \) are the farfield stresses, \( A \) and \( l \) are the amplitude and the wavelength of the perturbations, \( E, \nu \) and \( \mu \) are Young Modulus, Poisson ratio, and the friction coefficient, respectively, \( k \) is the farfield stress ratio, \( \sigma_{xz} / \bar{\sigma}_{xz} \), and \( U \) is the displacement along the fault.

\( (U=0 \text{ in the case of Fig. 10a). In the case presented in Fig. 10, the farfield stresses are defined by the vertical and the horizontal stresses.} \)

The host rock yields following a Coulomb failure criterion (Fig. 10b). Failure occurs when the difference between the principal stresses exceeds the internal frictional stress. This criterion is

\[ \sigma_1 > \sigma_2 + 2\mu \sigma_2 [(1+\mu^2)^{1/2} + \mu] \quad (B4) \]

where \( \sigma_1 \) and \( \sigma_2 \) are the maximum and minimum principal stresses, respectively (Jaeger et al., 2007, eq. 3.31). The stresses are calculated by solving eqs. B1-B3 as a function of space and the given fault displacement \( U \) with prescribed farfield stresses. All calculation assume \( \mu=0.7 \).

Appendix C

The flow model used for the calculations of the optimal wavelength of bumps generated as a shear flow instability is based on the work on folds and other instabilities arising from stiffness contrasts (Johnson and Fletcher, 1994; Smith, 1977). We follow the method and notation of Smith (1977) here with appropriate modifications made for the simple shear rather than pure shear geometry.
For a Newtonian fluid with negligible inertia, combining the momentum and continuity equations results in the biharmonic equation

\[ \nabla^4 \psi = 0 \] (C1)

where \( \psi \) is the stream function. The velocity field of the fluid is \( u = \frac{\partial \psi}{\partial y} \) and \( v = -\frac{\partial \psi}{\partial x} \). where \( u \) is the flow velocity in the x direction and \( v \) is the flow velocity in the y direction. The linear stability of the system is investigated by assuming a separable solution to \( \psi \) of the form

\[ \psi(x, y, t) = \varphi(y) \exp[iax]\exp[i\gamma t] \]

where \( \varphi(y) \) is an appropriate function of \( y \), \( \gamma \) is the growth rate, \( a = 2\pi / \lambda \) is the wavenumber and \( \lambda \) is the wavelength. Surface tension is neglected.

Substituting \( \psi \) into Eq. C1 results in

\[ \varphi''' - 2a^2 \varphi'' + a^4 \varphi = 0 \] (C2)

where primes are derivatives with respect to \( y \).

For power law fluids the rheology is \( \sigma'' \propto \dot{\varepsilon} \) where \( \sigma \) and \( \dot{\varepsilon} \) are the stress and strain rate tensors and \( n \) is a material constant. Smith (1977) performs a perturbation on the rheology as well as the deformation field for a basic flow of horizontal compression or extension. He shows that the linear stability equations are similar to the Newtonian fluid case. We extend the analysis for a basic flow of simple shear.

An appropriate form of \( \varphi(y) \) that respects the boundary conditions must be chosen. For a non-Newtonian flow, \( \varphi(y) = \sum_{i=1}^{4} A_i \exp[l_i y] \) where \( l_i \) are the four values of \( \pm a \sqrt{W} \pm \sqrt{W^2 - 1} \) for each layer with \( W = 2n - 1 \) (Smith, 1977).
With this choice, the boundary conditions can be expressed in terms of $\varphi(y)$. The coordinate system is chosen such that the bottom of lower layer (subscripted 1) is $y = 0$. The thickness of the lower layer is $h$ and the top layer is infinite. There is no slip and a continuity of traction between the layers, so the boundary conditions at $y = h$ are

$$\varphi_1 = \varphi_2$$  
(C3)

$$\varphi_1' = \varphi_2'$$  
(C4)

$$-(\varphi_1^{"} + a^2 \varphi_1) + m(\varphi_2^{"} + a^2 \varphi_2) = \frac{4}{\gamma} (m - 1) a^2 \varphi_1 E_{11}$$  
(C5)

$$-(\varphi_1^{"} - (2W_1 + 1)a^2 \varphi_1')/n_1 - m(\varphi_2^{"} - (2W_2 + 1)a^2 \varphi_2')/n_2 = -\frac{4(m/n_2 - 1/n_1)a^2 \varphi_1'}{\gamma} E_{12}$$  
(C6)

where the subscripts of 1 and 2 denote the bottom and top layers, respectively, $n$ are the power law exponents of the constitutive law, $m$ is the ratio of the effective viscosities at the basic state ($m = \eta_2 / \eta_1$ where $\eta_2$ and $\eta_1$ are the effective viscosities in the absence of the perturbation) and $\gamma$ is the growth rate normalized by the background shear strain. Note that the subscript convention is reversed from Smith (1977). The layer parallel compressive background strain is specified by $E_{11}$ and the simple shear strain by $E_{12}$. In this paper $E_{11}=0$ and $E_{12}>0$. This background flow is the major difference from previous work on boudinages which require that $E_{12}=0$ and $E_{11}<0$.

The no slip condition at the bottom of the lower boundary is achieved at $y = 0$ with the equations

$$\varphi_1 = 0$$  
(C7)

$$\varphi_1' = 0$$  
(C8)
The solution for the growth rate $\gamma$ as a function of wavenumber $a$ is found by posing the boundary conditions as a matrix system $MG=0$ where $G$ is the vector of constant coefficients $(A_1, A_2, A_3$ and $A_4$ for each layer). The top layer is assumed to be effectively infinite as it is much thicker than the bottom layer so $A_1 = A_2 = 0$ in that layer (layer 2).

Solving the equation $Det(M) = 0$ provides a solution for the growth rate $\gamma$ as a function of wavenumber $a$. We solve this equation numerically for a range of values of $m$, $n_1$ and $n_2$ in Fig. 11. We then select the optimum wavelength as that with the maximum growth rate and compare it to the observed wavelength of the bumps.
Figure 1. Geological maps and pictures of the Flowers Pit fault. **a)** Section of the geological map of Southwest Oregon showing the regional geological units (Jenks et al., 2007). MP-Miocene to Pliocene basalt and andesite. Q- quaternary sedimentary rock. P- Pliocene volcanic flows. A- Alluvium and covered area. Inset in the upper left shows the regional location in the context of the western U.S. and the square on the geological map marks the Flowers Pit general locale. **b)** The exposed (black lines) and the partly exposed (dashed lines) faults in Flowers Pit, marked on a Google Earth air photo. The Flowers Pit is located in the southwest part of a horst structure.
exposing normal faults that is striking northwest. Three large fault segments are marked by letters (SE: southeast segment, M: middle segment, NW: northwest segment). Photo and LiDAR imaging of the Southeast section of Flowers Pit Fault showing the continuous exposure of the surfaces. The image in (d) is produced by scanning the surface with ~2 million measured points with spacing of 0.5 cm.
Figure 2. Fault surface geometry in three exposures. a) Bumpy zone in the middle segment of the Flowers Pit fault. The main slip surface (Layer I) is underlain by a cohesive granular layer (Layer II), and by non-cohesive damage zone (Layer III). b) LiDAR data of fault surface.
1. Topography of the area inside the rectangle in (a) as a color-scale map rotated so that the X-Y plane is the best-fit plane to the surface and the mean striae are parallel to Y. Shown are two protruding bumps and one elongated depression with lengths of ~20 m and widths of ~2 m.

c) Large bump in the northern segment of Flowers Pit fault. d) Bumpy surface of normal fault in central Apennines, Italy. Note that in the three examples, in the direction parallel to slip, the fault surfaces are polished at scales of less than 5 m and bumpy at larger scales.
Figure 3. Fault surface topography and roughness. a) Maps of protruding bumps and profiles of the surface geometry parallel to slip. left: sample from the middle segment and right: sample from the northwest segment. The profiles show the faults are polished at scales < 5 m and bumpy at larger scales in the slip direction. b) Power spectral density calculated from six different fault
sections that have been scanned parallel to the slip orientation using ground-based LiDAR (upper
curves), and from two hand samples scanned by a profilometer in the lab (lower blue curves).
Each curve includes 300-600 individual profiles from the best preserved sections of the fault.
Dashed orange lines represent the noise level as established by scans of smooth, planar reference
surfaces for the Lidar (upper) and for the profilometer (bottom). A majority of the roughness in
the smooth hand samples is probably contributed by micro-erosion in layer I.
Figure 4. Damage characteristic of the Flowers Pit Fault from the exposed surface (top) toward the footwall (bottom) described by schematic stratigraphic column. The term ‘layer’ is used here to describe the damage as the fault zone displays structure that is relatively continuous parallel to the fault surface, but abruptly changes as function of the distance from the surface. The lengths of the boxes qualitatively indicate the rock cohesiveness. I: Band of slip surfaces (0.1-20 mm); II: Cohesive granular layer (0.05-1 m); III: Non-cohesive fragmented layer (0.3-2 m); IV: Non-systematic joints and fractures (~10 m); V: Jointed zone (>10 m).
Figure 5. Photographs of field exposures and thin sections of the slip band zone (Layer I) and the upper part of the cohesive granular layer (Layer II). a) A photo of slip surface shows the
extremely polished Layer I above the cohesive Layer II. The pen point is on fine material from Layer I which is injected to Layer II. The photo is rotated so that the fault surface is horizontal. b) Scanning Electron Microscope (SEM) image of grains on slip zone in Layer I. c) Layer I & II structure with key points labeled: 1) Typical sharp slip surface (100 microns wide). 2) Below the slip surface is Layer II which contains rounded aggregates of plagioclase and single crystals fragments. 3) Fracture which contains grains from the slip band and penetrates Layer II. 4) Zone of small-scale particles generated by the intrusion of Layer I grains. d) Close-in view of evidence for opening modes involving fluid flow. Feature labeled (1) is the slip surface. The arrow indicates a vein inside the branching fracture. e) Thick zone of Layer I showing several subparallel slip surfaces underlain by a fluidized ductile shear zone. f) Fluidized deformation with wavy layers (labeled 1) and boudins (labeled 2). Subfigures e & f are continuous in the sample.
Figure 6. The variations of the width of Layer II under protrusions and depressions of the fault surface. a) Variations of Layer II with in the exposure of the middle part of Flowers Pit Fault surface. The biggest bump (marked by B) overlies a the most thickened Layer II with a width >1 m at the tip of the arrow. The largest depression (marked by D) overlies the most thinned Layer II with a minimum width of 5 centimeters. Smaller perturbations of the width of Layer II are also observed under smaller protrusions and depressions. b) Two examples of local thickening of Layer II. Under protrusions (left) while thin appearance of the layer is observed under depressions (right). c) Maximum observed width of Layer II under large bumps measured from six protrusions, and minimum width of Layer II measured under ten large depressions. The background values measured in areas with relatively small amplitude variations of the surface (32 measurements). Error bars indicate 1 standard deviation.
Figure 7. Thick exposure of Layer II below the main fault surface. A clear sharp difference between the cohesive appearance of Layer II and the non-cohesive Layer III is observed. a) Small normal faults with a few centimeters of slip (white arrows) appear in Layer II. A sharply localized secondary fault parallel to the main one is exposed between Layer II and Layer III. b) Similar example which demonstrates the development of boudins in Layer II under sharp main slip surface. The interpretation in the text suggests that boudinage and decapitation of the bumps generates the observed structure.
Figure 8. Thin-section from Layer III shows plagioclase rich andesite-hosted shear bands (yellow arrows) and fractures (red arrows).
Figure 9. The density of fractures in the Flowers Pit Fault. Density is measured as total fracture length per area as a function of the distance from the fault. Blue diamonds: density values are calculated by mapping fractures in an area with a gridded frame and photographs. Beyond Layer III fracture density decays roughly as the square root of distance (black line). The two red squares are values of fracture density measured in four thin sections (such as in Fig. 7) from two different locations in Layer III. In these measurements, shear bands counted as fractures and their internal deformation is neglected. Brown circle represents the density of systematic large open cross-layer joints.
Figure 10. Measurements of six different faults in the fault zone (including the main fault) suggest a correlation between the average thickness of the cohesive layers and total slip. The rate of thickening is strongly reduced as a function of the total slip. In most of the exposures on the main fault the cohesive layer thickness is 10-20 cm. However, under protruding bumps (right figure) the layer is always thickened and the width locally can exceed one meter (red rectangle). The photo above shows the layer in a fault that is displaced 3 m.
Figure 11. Modeled stress field due to slip on a wavy fault. a) Calculated orientation and values of $\sigma_1$ following Chester and Chester (2000) model (Appendix B). We used our topography measurements (Fig 2, 3) to define a typical asperity aspect ratio, $H/L = 5 \times 10^{-3}$, and used a realistic non-critical stress ratio $\sigma_H/\sigma_V=0.5$. Inset: Modeled fault geometry. b) Critical slip distance on wavy slip surface as a function of asperity ratio, $H/L$, for three different stress ratios. A larger amount of slip before internal yielding is predicted for smoother surfaces.
Figure 12. Growth rate of bumps (asperities) as a function of wavelength normalized by layer thickness. Calculations are for a range of effective viscosity ratios ($m$) and stress-strain power law exponents ($n$) in each layer (subscripts 1 and 2). (See Appendix C). Growth rate is reported for a background strain of $E_{12}=1$ and scales linearly with the background strain (Appendix C). The wavelength with the peak growth rate for a given set of parameters dominates the flow and, thus, the preserved outcrop. Gray bar at the top of the figure indicates the range of observed wavelength/thickness ratios. The maxima of all of the computed curves are consistent with the range of observed bump wavelengths.